Effect of permafrost on the formation of soil organic carbon pools and their physical–chemical properties in the Eastern Swiss Alps

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Abstract: Current climatic conditions and the occurrence of discontinuous and sporadic permafrost in the Alps result in a low turnover rate and therefore accumulation of organic matter (OM) in soils. Alpine soils are thus highly sensitive to global warming that potentially promotes the mineralisation of soil organic matter (SOM). This process might increase the release of CO to the atmosphere. Our aim was to investigate the potential effect of permafrost thawing by the analysis of the physical–chemical soil properties of permafrost versus non-permafrost sites. Specifically, we i) quantified the SOM stocks at such sites, ii) characterised SOM and its physical and chemical fractions and iii) estimated the age range of the bulk soil and stable C-fraction (radiocarbon dating). In south-eastern Switzerland, two areas above the timberline and one below the timberline (where isolated permafrost was verified) were investigated in detail. At each site, the experimental set-up consisted in the comparison of nearby soils that were either influenced or not by permafrost. The C-stocks (down to the C horizon or rock surface) did not show a significant difference between permafrost and non-permafrost soils and were in the same range of 10–15 kg/m² in alpine (grassland) and subalpine (forest) sites. Above the timberline, the bulk SOM showed a distinct higher age at permafrost sites compared to non-permafrost sites. This higher age was even more evident in the stable C-fraction (resistant to an H O treatment), where ages of up to 11 ky in permafrost soils were recorded. The highest age obtained in the stable C-fraction in non-permafrost soils was around 4 ky. Consequently, climatic conditions and the occurrence of discontinuous permafrost resulted in a very low turnover rate of SOM. At the subalpine site, the difference between permafrost and non-permafrost sites was less. At both sites (alpine and subalpine), DRIFT (Diffuse Reflection Infrared Fourier Transform) was used to determine the functional groups in the bulk soil and in the stable C-fraction. In general, the stable C-fraction had a different composition compared to the bulk SOM at non-permafrost sites; this was mostly not the case at the permafrost sites. This confirms that different decomposition processes occur between permafrost and non-permafrost sites. Furthermore, permafrost sites accumulated more the low-density physical fractions of SOM that are potentially easily degradable. The obtained results suggest that a warmer climate may not necessarily lead to an increased CO release from SOM-degradation in permafrost soils compared to non-permafrost soils. High-alpine soils and OM furthermore integrate a multi-facetted response to the past and ongoing surrounding conditions. The melting of permafrost will most likely enhance vegetation growth, which to a certain degree will probably compensate for carbon losses on the long-term.

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Effect of permafrost on the formation of soil organic carbon pools and their physical-chemical properties in the Eastern Swiss Alps

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Abstract

Current climatic conditions and the occurrence of discontinuous and sporadic permafrost in the Alps result in a low turnover rate and therefore accumulation of organic matter (OM) in soils. Alpine soils are thus highly sensitive to global warming that potentially promotes the mineralisation of soil organic matter (SOM). This process might increase the release of CO_2 to the atmosphere. Our aim was to investigate the potential effect of permafrost thawing by the analysis of the physical-chemical soil properties of permafrost versus non-permafrost sites. Specifically, we i) quantified the SOM stocks at such sites, ii) characterised SOM and its physical and chemical fractions and iii) estimated the age range of the bulk soil and stable C-fraction (radiocarbon dating). In south-eastern Switzerland, two areas above the timberline and one below the timberline (where isolated permafrost was verified) were investigated in detail. At each site, the experimental set-up consisted in the comparison of nearby soils that were either influenced or not by permafrost. The C-stocks (down to the C horizon or rock surface) did not show a significant difference between permafrost and non-permafrost soils and were in the same range of 10 – 15 kg/m^2 in alpine (grassland) and subalpine (forest) sites. Above the timberline, the bulk SOM showed a distinct higher age at permafrost sites compared to non-permafrost sites. This higher age was even more evident in the stable C-fraction (resistant to an H_2O_2 treatment), where ages of up to 11 ky in permafrost soils were recorded. The highest age obtained in the stable C-fraction in non-permafrost soils was around 4 ky. Consequently, climatic conditions and the occurrence of discontinuous permafrost resulted in a very low turnover rate of SOM. At the subalpine site, the difference between permafrost and non-permafrost sites was less. At both sites (alpine and subalpine), DRIFT (Diffuse Reflection Infrared Fourier Transform) was used to determine the functional groups in the bulk soil and in the stable C-fraction. In general, the stable C-fraction had a different composition compared to the bulk SOM at non-permafrost sites; this was mostly not the case at the permafrost sites. This confirms that different decomposition processes occur between permafrost and non-permafrost

sites. Furthermore, permafrost sites accumulated more the low-density physical fractions of SOM that are potentially easily degradable.

The obtained results suggest that a warmer climate may not necessarily lead to an increased CO_2 release from SOM-degradation in permafrost soils compared to non-permafrost soils. High-alpine soils and OM furthermore integrate a multi-facetted response to the past and ongoing surrounding conditions. The melting of permafrost will most likely enhance vegetation growth, which to a certain degree will probably compensate for carbon losses on the long-term.

Keywords: Alpine soils, climate, stocks of soil organic carbon, stability of soil organic carbon, radiocarbon dating

52 **1. Introduction**

53 Permafrost is a widespread phenomenon in cold alpine and subarctic environments. Climate warming is
54 highly threatening the stability of permafrost soils in areas with mean annual air temperatures close to 0
55 °C. A thawing of permafrost in those sensitive ecosystems will not only enhance the risks regarding,
56 for example, slope stability; it will also increase the vulnerability of soil organic matter (SOM) to rapid
57 microbial decomposition that was previously stabilised by freezing temperatures (Kääb et al., 2007;
58 Schmidt et al., 2011).

59 Discontinuous permafrost distribution in the Alps is a function of altitude and exposure and can be
60 assumed to potentially occur at north-exposed sites above an altitude of 2400m asl, and on south-facing
61 sites above 3000m asl, respectively (Nötzli and Gruber, 2005). Due to the local microclimate, sporadic
62 permafrost can additionally exist below the timberline at very shaded sites (Kneisel et al., 2000). In the
63 Younger Dryas (late Pleistocene), permafrost in the south-eastern part of Switzerland had been present
64 in areas above 1950 m asl in north-exposed slopes, and above 2450 m asl on south-facing slopes (i.e.
65 500 m to 600 m below the present-day limits of discontinuous permafrost). The formation of Alpine
66 soils mostly started in the late Pleistocene or at the beginning of the Holocene and was characterised by
67 alternating warmer and cooler conditions. Some of these warmer inter-phases occurred during the
68 Older Atlanticum (5 to 8 ky BP) that is characterised as the 'climatic optimum' (Frauenfelder et al.,
69 2001; FOEN, 2005; Böhlert et al., 2011b). Permafrost distribution, soil genesis and SOM formation
70 was thus not always straightforward. Current soil properties integrate the effect of such climatic
71 variability.

72 A changing climate is generally assumed to affect soil organic carbon. What this might mean for soils
73 having a complex evolution is, however, largely unknown. The complex terrestrial carbon cycle is
74 typically controlled by three main factors: (i) climate (temperature and precipitation; Sanderman et al.,
75 2003) that is influencing vegetation establishment and microbial decomposition; (ii) the biochemical
76 composition of organic residues and SOM (Gleixner et al., 2001) which is highly dependent on plant

77 species and decomposition processes and (iii) the removal of organic carbon out of the carbon cycle —
78 or at least decelerated decomposition rates due to the stabilisation of organic matter, e.g. in organo-
79 mineral complexes (Krull et al., 2003).
80 The fractionation of SOM in its different decomposition and stability phases is essential because it
81 predetermines the vulnerability of SOM to environmental changes (Budge et al., 2011). Several
82 chemical and physical fractionation techniques exist to separate the labile C-fraction (only slightly
83 degraded OM) that mostly responds more rapidly to environmental changes, from the more stable,
84 recalcitrant mineral-associated C-fractions showing longer turnover times (Leifeld and Kögel-Knabner,
85 2005; Kögel-Knabner et al., 2008). The mechanisms involved in the stabilisation of the mineral-
86 associated soil carbon pools are still not entirely clear (von Lützow et al., 2006; Helfrich et al., 2007).
87 In permafrost environments mixing of soil layers due to cryoturbation can further influence the
88 formation of soil carbon pools: Freeze and thaw cycles enable the redistribution of organic carbon from
89 the surface to a greater soil depth or the other way around. The former process may retard its
90 decomposition and enables the soil to restart carbon accumulation in the topsoil layers (Ugolini et al.,
91 2006; Kaiser et al., 2007) whereas latter one will bring organic carbon to the surface horizons and
92 potentially may promote C-mineralisation.
93 Available data from alpine tundra soils show relatively large organic carbon concentrations in cold
94 alpine soils and a high abundance of less decomposed labile organic carbon with increasing altitude
95 (Jacot et al., 2000; Leifeld et al., 2009; Budge et al., 2011). It is generally assumed that permafrost soils
96 in arctic and subarctic lowland environments store a considerable amount of organic carbon, even
97 though possible mechanisms of carbon stabilisation by processes other than freezing are still unknown
98 (Hugelius and Kuhry, 2009; Schmidt et al., 2011). Most of the studies were done in high northern
99 latitudes with arctic and subarctic ecosystems (e.g. Bird et al., 2002; Zimov et al., 2006; Schuur et al.,
100 2008; Hugelius and Kuhry, 2009). Anoxic wetland conditions are typical for northern permafrost soils
101 and are responsible for the high accumulation of organic carbon. These specific processes cannot

directly be transferred to alpine permafrost soils that may differ in, for example, soil genesis, hydrology and topography. Furthermore, only little data is available about the storage of organic carbon, its chemical characteristics and turnover times in permafrost soils of the Alps. Consequently, the present work focuses precisely on these issues by comparing permafrost soils with nearby non-permafrost soils at the alpine and subalpine range in the south-eastern Swiss Alps.

The following research questions arose: (i) Are the SOM stocks at permafrost-sites really higher compared to non-permafrost sites? (ii) Are the SOM characteristics and ages of C-fractions at such sites different between permafrost and non-permafrost soils? (iii) Do the SOM stocks and chemical characteristics differ between alpine permafrost and subalpine permafrost soils?

Our hypotheses for the study were: (i) permafrost soils contain a larger amount of labile organic matter than non-permafrost soils, (ii) permafrost soils differ in their SOM quality and have a higher mean residence time compared to non-permafrost soils, (iii) the subalpine permafrost sites experience a warmer climate compared to alpine permafrost sites which should affect C-storage and C-chemistry.

2. Materials and methods

2.1. Study area

The study areas are located in the Upper Engadine (Switzerland) (Fig. 1). The Lateglacial and Holocene history of the Upper Engadine area is documented in, e.g., Suter (1981) and Böhlert et al. (2011b). The soils are skeleton-rich Leptosols (humic-skeletic) to Cambisols (WRB, 2006). The glacial till deposited in these areas consists of granite/gneiss ('Julier Granite', Table 1). According to the Soil Taxonomy (Soil Survey Staff, 2010), the soil moisture regime is udic (humid conditions, < 90 days/ per year having a dry soil) at all sites and the soil temperature regime is cryic (mean annual temperature < 8 °C). Maximum precipitation occurs during the summer and autumn months. Two alpine and one subalpine site were chosen (Fig. 1, Table 2):

1. Alpine areas (Bever and Albula):

These sites are situated at approximately 2700m asl. The climate is characterised by a mean annual temperature of -2.6 °C and a mean annual precipitation of >1000mm and according to the Köppen-Geiger classification defined as ET-climate (Kottek et al., 2006). At the Bever site, the occurrence of permafrost was measured using geoelectrical sounding. At both alpine sites the permafrost distribution was defined using an alpine permafrost map (Böckli et al., 2012). Permafrost soils are favoured at north-exposed slopes, whereas permafrost-free sites are found at south-facing sites. At the Bever site, permafrost features on the soil surface were further indicating cryoturbation caused by freeze and thaw cycles. The vegetation cover was predominantly grasses and herbaceous plants (Caricetum curvulae, Geo montani-Nardetum) and was usually denser at the permafrost-free sites, especially at the Bever site.

2. Subalpine area (Spinas):

This site is located below the timberline at 1800 m asl (at present the timberline is between 2200m and 2300m asl.), and represents one of the special places where isolated permafrost lenses could be confirmed by several geophysical techniques also at lower altitudes (Kneisel and Hauck, 2003; Kneisel and Schwindt, 2008). Mean annual temperature is 1°C and the mean annual precipitation is 1050 mm (Schwarb et al., 2000). According to the Köppen-Geiger classification, the climate is defined as Dfc (Kottek et al. 2006). Permafrost lenses and nearby sites without permafrost could both be found at north-exposed scree slopes due to differences in the microrelief and therefore site-specific climatology. There was no difference in the vegetation cover between the permafrost and non-permafrost soils. The forest is a Larici-Pinetum cembrae, with blueberry (*Vaccinium myrtillus*) and cranberry (*Vaccinium vitis-idaea*) in the herbaceous layer.

2.2. Experimental set-up

151 At each site (Bever, Spinas and Albula) 6 soil profiles (half of them having permafrost and half of them
152 without permafrost) were excavated. Approximately two to four kilograms of soil material (Hitz et al.,
153 2002) was collected per soil horizon. Soil samples down to the C horizon were taken. Soil bulk density
154 was determined by a soil core sampler. At the subalpine site Spinas, an additional set of 10 profiles
155 (down to a depth of 20 to 30 cm) was sampled to enlarge the database particularly with respect to soil
156 chemical properties. Half of these soils had permafrost and half were without permafrost.
157
158 *2.3. Soil Chemistry and Physics*
159 Total C and N contents of the soil were measured with a C/H/N analyser (Elementar Vario EL). Soil
160 pH (in 0.01 M $CaCl_2$) was determined on air-dried fine earth samples using a soil:solution ratio of
161 1:2.5. As the soils did not contain inorganic carbon (Julier granite is the parent material; cf. also Table
162 1), the total C contents equals the organic carbon content.
163 After a pre-treatment of the samples with H_2O_2 (3%), particle size distribution of the soils was
164 measured using a combined method consisting of sieving the coarser particles (2000 - 32 µm) and the
165 measurement of the finer particles (< 32 µm) by means of an X-ray sedimentometer (SediGraph 5100).
166
167 *2.4. Chemical fractionation of organic matter*
168 Our conceptual approach was based on the finding that partial oxidative degradation of OM leaves
169 behind intrinsically resistant as well as mineral-protected organic materials. This method is based on
170 the oxidation of OM by 10% H_2O_2 (Plante et al., 2004, modified; Eusterhues et al., 2005). 1 gram of
171 air-dried, untreated soil (< 2mm) was wetted for 10 min with distilled water in a 150-ml beaker.
172 Afterwards, 90 ml of 10% H_2O_2 were added. The procedure was run at a temperature of 50 °C and in a
173 closed systems throughout the treatment period. Peroxide treatments were performed for 168h (7 days).
174 At the end of the treatment the samples were washed three times with 40 ml deionised water and
175 freeze-dried, their weights recorded, analysed for total C and N content and ^{14}C dated. Functional

176 groups were determined using DRIFT (Diffuse Reflection Infrared Fourier Transform). The amount of
177 organic C after the H_2O_2 treatment was related to the initial content using a mass-balance approach to
178 obtain the corresponding recoveries with

179 $$\text{recovery of stable C (\%)} = (gC_{after} / gC_{before}) \times 100 \quad (1)$$

180 Recovery values were calculated also for nitrogen.

181 In the following, the OM that resisted to the H_2O_2 treatment is called the 'stable C'. The soil fraction
182 oxidised by H_2O_2 is the 'labile C'.

183

184 *2.5. Physical fractionation of organic matter*

185 The samples were air-dried. Large aggregates were gently broken by hand and sieved to < 2 mm. The
186 density fractionation procedure was adapted from Christensen (1992). Carbon fractions were separated
187 according to the following densities: <1.0 g cm^{-3} (without ultrasonic treatment) 1.0-1.6 g cm^{-3} (without
188 ultrasonic treatment), 1.0-1.6 g cm^{-3} (with ultrasonic treatment) 1.6-2.0 g cm^{-3} (with ultrasonic
189 treatment) and >2 g cm^{-3} (with ultrasonic treatment). The lightest fractions with a density <1 and 1-1.6
190 g cm^{-3} (without ultrasonic treatment, both fPOM) include the fresh or only slightly degraded organic
191 matter, but also inherent recalcitrant material such as charred residues. The fraction with a density 1-1.6
192 g cm^{-3} (with ultrasonic treatment, oPOM) represents carbon that is occluded within aggregates and
193 consequently physically protected for degradation. The fractions 1.6-2 g cm^{-3} and >2 g cm^{-3} (with
194 ultrasonic treatment, both mOM) are considered to be the more mineral associated C-fractions and
195 therefore chemically protected through interactions with mineral surfaces (von Lützow et al., 2006).
196 Forty grams of fine earth was added to 400 mL of deionised water, slightly stirred for 12 h, then
197 centrifuged for 10 min at 5000 g. The supernatant was sucked off and evaporated, so obtaining the <1 g
198 cm^{-3} fraction. The residual sample was dried at 40 °C and mixed to a sodium polytungstate solution
199 with a density of 1.6 g cm^{-3} to separate the 1.0-1.6 g cm^{-3} fraction from the >1.6 g cm^{-3} one. The
200 residual sample was again dried at 40 °C, mixed to a sodium polytungstate solution with a density of

1.6 g cm^{-3} but now an ultrasonic treatment of 50 J mL^{-1} for 40 s was applied to the suspension, which then was centrifuged for 30 min at 3000 g. The supernatant was sucked off and evaporated, so obtaining the 1.0-1.6 g cm^{-3} fraction. The same procedure was finally applied to the residue using a polytungstate solution with a density of 2.0 g cm^{-3} so as to separate the 1.6-2.0 g cm^{-3} from the >2 g cm^{-3} one. All of the fractions were washed three times with deionised water, dried at 40 °C and analysed for total C and N content.

2.6. DRIFT measurements

Relative peak intensities were used for DRIFT analysis (Bruker, Tensor 27). Spectra were recorded from 4000 to 250 cm^{-1} using a powder containing 9-12 mg of sample (3% of the total weight) and 291-388 mg KBr (97% of the total weight). The samples were homogenised in a mill using a fine ball-mill (Zr) for 45 s (frequency 25.0). Prior to measurement, the samples were dried in the oven at 70°C.

In order to quantify the relative changes in the FT-IR spectra, we divided the values of the relative intensity (area) of each peak by the sum of the relative intensity and multiplied it by 100 using the software OPUS 6. For the following peak range an individual peak search and calculation of relative intensities (area) was done: (base1/base2)(cm^{-1}): 2980/2880, 1725/1710, 1660/1630, 1620/1600, 1535/1500, 1495/1470, 1470/1440, 1413/1333, 1190/1127, 1116/1050 and 1080/1030. The integration method using a linear background between the found bases (individual samples) and absorbance values was taken to calculate the relative concentration (OPUS 6). Major IR absorption bands and functional groups assignments are given in Table 3. Aliphatic compounds were calculated using the IR range 1470-1440 cm^{-1}. The range between 2980-2828 cm^{-1} was not considered because water may absorb in this range and thus disturb the calculation procedure. DRIFT measurements were performed on the H_2O_2 treated (stable C) and untreated soil material (bulk soil).

2.7. Carbon and nitrogen stocks

226 The total soil organic carbon (SOC) and stable carbon stocks were calculated down to the C horizon or
227 rock surface according to the following equation:

228 $$S_{stock} = \sum_{i}^{dz} C_i d_i \rho_i (1 - RM) \quad (2)$$

229 where S_{stock} denotes the abundance (kg/m^2) of the corresponding element and fraction, C the
230 concentration of the element in the corresponding fraction (kg/t), d_i the thickness of layer i (m), ρ = soil
231 density (t/m^3) and RM the mass proportion of rock fragments/soil skeleton.

232

233 *2.8. Radiocarbon dating of organic matter fractions*

234 Samples were cleaned using an acid-alkali-acid (AAA) treatment. The samples were then heated under
235 vacuum in quartz tubes with CuO (oxygen source) to remove any absorbed CO_2 in the CuO. The tubes
236 were then evacuated, sealed and heated in the oven at 900°C to obtain CO_2. The CO_2 of the combusted
237 sample was mixed with H_2 (1:2.5) and catalytically reduced over iron powder at 535°C to elemental
238 carbon (graphite). After reduction, the mixture was pressed into a target and carbon ratios were
239 measured by Accelerator Mass Spectrometry (AMS) using the tandem accelerator of the Laboratory of
240 Ion Beam Physics at the Swiss Federal Institute of Technology Zurich (ETHZ). The calendar ages were
241 obtained using the OxCal 4.1 calibration program (Bronk Ramsey, 2001; 2009) based on the IntCal 09
242 calibration curve (Reimer et al., 2009). Calibrated ages are given in the 2σ range (minimum and
243 maximum value for each).

244 Percent modern carbon (pMC) was calculated for obtaining the age of the labile organic matter
245 fraction:

246 $$pMC = 100 \times \left(e^{-\frac{\left(t + y - 1950 / 1.03\right)}{8033}} \right) \quad (3)$$

247 The pMC of H_2O_2-removable organic carbon (OC) was calculated from mixing models (Mikutta et al.,
248 2006):

$$I - C_L = \frac{\left((I - C_U) \cdot UOC - (I - C_S) \cdot SOC\right)}{LOC}$$

249 (4)

250 where $I - C_U$, $I - C_L$ and $I - C_S$ are, respectively, the values of untreated, H_2O_2 removable and H_2O_2
251 resistant (stable) OM. *UOC (untreated OC)*, *LOC (labile OC)* and *SOC (stable OC)*, correspond to the
252 respective OC contents.
253
254 *2.9. Statistics*
255 The normal distribution of the data was verified with a Shapiro-Wilk test. A T-Test for data showing
256 homogeneity of variance (F-Test) and Welch-Test for data not showing homogeneity of variance was
257 performed in order to test the difference between two groups. For all tests a level of significance of
258 0.05 was considered. Statistical analyses were performed using the statistical software R (R
259 development core team 2008, RVersion 0.95.263).
260
261 **3. Results**
262
263 *3.1. General soil properties, SOM stocks and chemical fractions*
264 All soils were acidic and the pH ranged from about 3.8 in the topsoil up to 4.6 in the deeper soil
265 horizons (Table 4). The lower horizons and parent material consisted of loamy sand to sandy loam,
266 whereas in the topsoil a higher silt and clay content could be measured giving rise to a sandy loam, a
267 sandy clay loam or even a silt clay loam.
268 At the alpine and subalpine permafrost sites, average C-concentrations of 17±3 g/kg (Bever; 0 – 30 cm
269 depth), 54±28 g/kg (Albula; 0 – 20cm depth) and 141±88 g/kg (Spinas; 0 – 30cm depth), respectively,
270 were measured in the bulk-topsoils. Non-permafrost soils at the Bever site showed a significant higher
271 C-concentration (110±30 g/kg) compared to the permafrost site, whereas the concentration at the
272 Albula non-permafrost site (47±20 g/kg) was similar to the permafrost soil. The subalpine site (Spinas)

273 indicated a (non-significant) trend towards a lower C-concentration in the permafrost soils (67±29
274 g/kg) compared to the subalpine permafrost soils. With the exception of the Bever site, there was no
275 significant difference in the carbon concentration between permafrost and non-permafrost soils ($p < 0.05$).
276
277 After oxidation of the labile organic matter by H_2O_2, the median stable C-recovery value for the alpine
278 area was near 13% for the permafrost soil (PF) and 9% for the non-permafrost soil (nPF). The
279 subalpine area had median recovery values of 9% (PF) and 10% (nPF), respectively (Fig. 2). In the
280 subsoils, where less labile organic matter is present, higher recoveries of the stable C-fraction were
281 measured (Table 4).
282 As found in previous investigations in the Alps (Favilli et al., 2008; Egli et al., 2010), the C/N ratio of
283 SOM dropped abruptly after the H_2O_2-treatment (Fig. 2). The median C/N-value of the untreated soils
284 was near 9 (PF) and 10 (nPF) for the alpine area and 18 (PF) and 13 (nPF) for the subalpine area. After
285 the H_2O_2-treatment, the C/N ratio (median) was only 2 (PF) and 1 (nPF) for the alpine site and 5 (PF)
286 and 4 (nPF) for the subalpine area. Compared to the untreated soils, the H_2O_2-treatment thus leads to a
287 relative enrichment of nitrogen. In general, the C/N values obtained for the stable C-fraction (H_2O_2-
288 resistant) are quite low. The C/N ratio of the labile C-fraction was consequently higher (Table 4). This
289 fraction is relatively enriched in carbon.
290 The calculated OM-stocks for the bulk soil, labile and stable C-fractions are given in Figure 3. At both
291 altitudes no distinct differences between permafrost and non-permafrost sites were detectable. For both
292 type of sites (alpine and subalpine), stocks of about 10 kg/m^2 for the permafrost and 15 kg/m^2 for the
293 non-permafrost soils were obtained.
294
295 *3.2. Soil organic matter characteristics*
296 The DRIFT spectra of the bulk SOM and its stable C-fractions (oxidised by H_2O_2) deliver further
297 information about its chemical characteristics.

298 In both areas (alpine and subalpine sites) the topsoil bulk organic matter consisted mostly of aliphatic
299 groups (data not shown), phenolic functional groups and, to a lesser extent, of polysaccharides and
300 secondary alcohols (Fig. 4). Alpine permafrost bulk soil was richer in secondary alcohols and phenolic
301 groups but showed a lower percentage of polysaccharides compared to the adjacent non-permafrost
302 bulk soil. A slightly higher abundance of secondary alcohols and polysaccharides but a lower
303 proportion of phenolic groups was measured in the subalpine permafrost bulk soil than in the alpine
304 permafrost bulk soil. Humification processes are related to the preferential oxidation of plant
305 polysaccharides and phenolic hydroxyl groups and are further characterised by an enrichment of
306 secondary alcohols and N-containing compounds in the stable C-fraction (Fig. 4). The relative
307 enrichment or depletion was constantly lower in the permafrost soils (alpine and subalpine sites),
308 pointing towards differences in the decomposition process when compared to the non-permafrost soils.
309
310 *3.3. Soil organic matter and density fractionation*
311 An overview of the C-concentrations (g/kg soil) of the different density fractions in the subalpine soils
312 is given in Figure 5. Different carbon concentrations were obtained in the individual density fractions
313 for permafrost and non-permafrost soils (subalpine site). The largest differences were measured in the
314 light C-fractions fPOM (<1 and 1 – 1.6 g/cm^3) and oPOM where permafrost soils contained about 59%
315 fPOM and 27% oPOM of the total carbon, respectively. At permafrost sites, the recovery of C was
316 constantly decreasing with increasing density (from 1 – 1.6 to >2 g/cm^3). Consequently, the lowest
317 percentages of carbon were detected in the mineral-associated fractions (mOM), showing values of
318 about 12% (1.6 – 2.0 g/cm^3) and 2% (>2.0 g/cm^3). In contrast, non-permafrost soils had only 42% of
319 fPOM and 16% of oPOM. In the non-permafrost soils, the mineral-associated C-fractions contributed
320 to a greater extent to the total carbon recovery with values of about 32% (1.6 – 2.0 g/cm^3) and 10%
321 (>2.0 g/cm^3), respectively. The C/N ratio decreased in both soils with increasing density, from mean
322 values of 28 and 26 (fPOM: <1 g/cm^3) in permafrost and non-permafrost soils, respectively, down to

15 in permafrost and 12 in non-permafrost soils in the heaviest soil fraction (>2 g/cm^3) (data not shown).

3.4. Radiocarbon dating

Carbon-14 analyses were done for the bulk soil and stable C-fractions in the upper soil horizons of the alpine and subalpine area. The age of these fractions are given in Table 5 for the alpine sites and in Table 6 for the subalpine sites. Above the timberline, the bulk SOM in permafrost soils had a higher age than the adjacent non-permafrost soils. Such differences were even more obvious in the stable C-fraction in the permafrost soils where ages up to 3 ky (0 – 10 cm) and 11 ky (10 – 30 cm) were measured, compared to a maximum age of 2.8 ky (0 – 10 cm), and approx. 4 ky (10 – 30cm) in non-permafrost soils.

Below the timberline, the age of the bulk SOM (0 – 15cm) was similar between permafrost and non-permafrost soils. In contrast, the stable C-fraction in permafrost soils (0 - 15 cm) was younger than in the non-permafrost soils with a maximum age of up to 1.3 ky, whereas the stable C-fraction at non-permafrost sites (0 - 15 cm) showed ages up to 3.6 ky. This pattern changed in the deeper soil horizons (15 – 30 cm), where higher ages in permafrost bulk soils (up to 0.5 ky) compared to non-permafrost bulk soils (modern age) were measured.

4. Discussion

4.1. SOC stocks and distribution

In the investigated alpine and subalpine soils, the stocks of bulk soil organic carbon (Fig. 3) was in the lower range of previously reported results (Egli et al., 2008; Favilli et al., 2008; Hagedorn et al., 2010). Hitz (2002) and Egli et al. (2008) found carbon stocks of up to more than 40 kg/m^2 in alpine soils with frequent values between 15 – 25 kg/m^2. Hagedorn et al. (2010) measured carbon stocks of 15 kg/m^2 for subalpine Swiss forest soils at similar altitudes. The variability in our data was, however, relatively

high. Carbon concentrations found in alpine and subalpine soils are consistent with values already reported for grassland and forest soils at those altitudes (Favilli et al., 2008; Egli et al., 2012).

Compared to our investigated soils, arctic and subarctic lowland permafrost soils usually have higher organic carbon stocks. Hugelius and Kuhry (2009), for example, reported organic carbon stocks of 38 kg/m^2 (>100cm soil depth) in taiga and tundra permafrost soils of northeastern European Russia. In northern circumpolar permafrost soils, Tarnocai et al. (2009) even found carbon abundances of 32.2 – 69.6 kg/m^2 (0 – 100 cm soil depth). Compiled data for arctic, subarctic and boreal Cryosols revealed carbon stocks of 28.5 kg/m^2, 62.6 kg/m^2 and 77.6 kg/m^2, respectively in northern Canada (Tarnocai et al., 2000). Wetland conditions and cryoturbation, particularly in permafrost soils, are major soil forming processes in arctic and alpine regions and may contribute to a higher long-term storage of C in soils (Ugolini et al., 2006; Kaiser et al., 2007).

Surprisingly, the alpine grassland and subalpine forest sites showed similar organic carbon stocks (in permafrost and non-permafrost soils) in our investigations areas (Fig. 3). A higher primary productivity and consequently higher litter production and C inputs in the root zone were expected for the forested sites. The lack of difference might be explained by a higher litter input in the alpine zones and its temperature-limited conservation until present time and/or by a high productivity of grassland ecosystems. An additional possibility could be that the C input at high altitudes (with soils actually having permafrost condition) was greater during a certain period of time due to warmer climate and a different vegetation type.

The higher C storage capacity in soils having grassland is supported by a meta-analysis of Guo and Gifford (2002). They reported an equivalent or even higher potential of grassland ecosystems to store organic carbon than forest soils. Additionally, Tate et al. (2000) showed higher C-stocks (+13%) in the New Zealand tussock grassland than in a mountain beech forest (19.9 vs 16.7 kg/m^2). Grasslands add organic matter in particular from below-ground (low shoot-to-root ratio) to the soil (Brown and Lugo, 1990). Hitz et al. (2001) even observed an increasing contribution of the dense fine root system to the

373 total carbon and nitrogen content and decreasing root turnover times in Alpine grasslands with
374 increasing elevation. Furthermore, Jackson et al. (2002) examined in a global analysis the relationship
375 of precipitation and to organic carbon storage in grassland and forest soils. They showed that grassland
376 productivity was strongly correlated to precipitation, with an increased total carbon content of around
377 43% compared to forest sites at precipitation rates of 1000 mm.
378 The climate experienced several variations during the late Pleistocene and Holocene and pedogenesis
379 has not been continuous and completely similar. A particularly warmer phase was recorded for the
380 Older and Younger Atlantic period (Maisch et al., 1999) when the treeline was 150-250 higher than the
381 present-day one (i.e. Burga, 1991; Favilli et al., 2010). Soil formation (chemical weathering) and the
382 storage mechanisms of organic C in the soil were consequently different to those of the present-day for
383 several millennia. This could mean that pedogenesis at the nowadays alpine sites occurred under a
384 different vegetation that had more similarities to the subalpine sites (more shrubs or even trees; Carnelli
385 et al., 2004). The uppermost tree limit reached 2500±100 m asl in the Valaisan Alps during the
386 Holocene (Carnelli et al., 2004); i.e. it was 250±100 m higher than today's potential tree line.
387 Consequently, the timberline would have reached 2400±100 m asl. The absence of differences between
388 alpine and subalpine soils could be due to the palaeo-environmental effect.
389
390 *4.2. SOC pools*
391 In Alpine environments the highest C and N concentrations are usually found in the labile C- and N-
392 fraction, whereas the stable C- and N-fractions are only a minor part of the total C and N (Egli et al.,
393 2012). The relative proportion of the labile C-fraction was relatively high, ranging between 87 – 91%,
394 and similar at all investigated soils (PF and nPF) of the alpine and subalpine sites. The median recovery
395 value of the stable carbon fraction was consequently quite low (median: 9 – 13%, Fig. 2) but consistent
396 with values reported by other authors who measured a stable carbon fraction in the range of about 10 –
397 20% in Alpine soils having ages even up to 17 ky (Favilli et al., 2008; Egli et al., 2009; Egli et al.,

398 2012). It seems that the permafrost soils at the subalpine sites contained more weakly degraded organic
399 matter than the alpine soils which gave rise to an increased C/N ratio (Table 4).
400 The stable C-fraction characterises an inaccessible, passive SOM pool and its stability is attributed to
401 natural compound recalcitrance deriving from its biochemical characteristics, such as the presence of
402 aliphatic macromolecules (lipids, cutans, algaenans, and suberans), charcoal and lignins and
403 particularly to various processes of interaction with mineral surfaces (von Lützow et al., 2006).
404 Compared to the bulk soil, we measured higher ^{14}C-ages in the stable C-fraction (Tables 5 and 6), a
405 decrease of OH-deformation and C-O stretching of phenolic groups, an increase in the relative amount
406 of proteic amide functional groups (N-H stretching of proteic amides) (Fig. 4) and a consistent decrease
407 of the C/N ratio (Fig. 2). This is in agreement with recently reported results where it is was
408 hypothesised that protein-rich, most probably microbially processed, compounds have the capability to
409 strongly adsorb onto mineral surfaces and are therefore crucial for the stability of organo-mineral
410 particles (Kleber et al., 2007; Sollins et al., 2009). Also in the mineral-associated C-fractions (mOM:
411 1.6 – 2 and > 2.0 g/cm^3) a lower C/N ratio was measured, pointing towards a lower C-loading with
412 increasing particle density and an accumulation of proteinaceous material. Figure 4 furthermore gives
413 the impression that in permafrost soils (alpine and subalpine), partially less degraded OM was able to
414 attach directly to mineral surfaces (Fig. 4; see polysaccharides and phenolic groups).
415 The C-recovery after the H_2O_2-treatment was consistent and low in all investigated soils (see chapter
416 3.1), but was not correlated to the percentage of carbon found in the heavier mineral-associated C-
417 fractions (mOM: 1.6 – 2 and > 2.0 g/cm^3) where a C-recovery of up to 42% was measured (see chapter
418 3.3). Von Lützow et al. (2006), for example, reported that density fractionation makes only a rough
419 differentiation between an active and passive organic matter pool. While the light C-fraction is a good
420 representation of the active pool, the heavy C-fraction contributes to the intermediate and passive pool.
421 An explanation might be given by the model proposed by Kleber et al. (2007). They suggested a
422 layered structure of organic matter attachment on mineral surfaces. A strong and direct interaction with

423 the mineral surface is favoured in situations where protein-rich material is adsorbed and forms a stable
424 inner-layer zone ('contact zone'). Other layers, whose compounds are not directly attached to the
425 mineral surface, surround this contact zone. The compounds in the outer layer are only loosely bound
426 by, for example, cation bridging or hydrogen bonding, which leads to a shorter residence times for the
427 individual molecular fragments. The thickness of this outer region depends more on the material input
428 than on the availability of binding sites (Kleber et al., 2007). Accordingly, we hypothesise that the
429 chemical treatment (H_2O_2) was better suitable to isolate the stable inner contact zone whereas the
430 physical fractionation method (density fractionation) cannot separate layers of different bonding-
431 strength.
432
433 *4.3. SOM quality*
434 In alpine soils, the relatively small-scale heterogeneity of the vegetation (see chaper 2.1) reflected the
435 impact of the microtopographic position and microclimate (influence of snow cover, soil moisture etc;
436 Haugland and Beatty, 2005). The bulk SOM at alpine permafrost and alpine non-permafrost sites
437 differed slightly in its chemical composition (Fig. 4). In alpine permafrost bulk soils a higher
438 abundance of secondary alcohols, phenolic groups and proteic amides, but a lower percentage of
439 polysaccharides were measured compared to alpine non-permafrost bulk soils. Additionally, the stable
440 C-fraction in alpine permafrost soils was less enriched in those compounds. These findings indicate
441 differences either in plant species and/or decomposition processes between alpine permafrost and
442 alpine non-permafrost soils. It is well known that the amount of residue input depends on the
443 vegetation productivity, whereas SOC quantity and quality is largely determined by plant species
444 (Helfrich et al., 2006) and the extent of decomposition that is related to the microbial community
445 (Ohtonen et al., 1999; Kögel-Knabner et al., 2002). We hypothesise therefore that the differences in the
446 chemical composition of SOM between alpine permafrost and alpine non-permafrost soils can be
447 related to a lower C input, a reduced microbial activity and associated decomposition process in alpine

448 permafrost soils. Differences in the decomposition conditions are further supported by the radiocarbon
449 dating of the alpine bulk soils and its stable C-fractions. ^{14}C-ages of bulk soils and related labile and
450 stable C-fractions (Table 5) were distinctly higher in alpine permafrost soils than in alpine non-
451 permafrost soils. This indicates a higher mean residence time and therefore a lower organic matter
452 turnover. Also Favilli et al. (2009) reported high ages of the resilient C-fraction, often up to 11-14 ky
453 (an in one case up to 17.5 ky; 4-8 cm depth) for topsoils in the subalpine region (close to the
454 timberline) and up to 11 ky for topsoils in alpine soils. Radiocarbon dating of the stable C-fraction can
455 also be used as a surface age indicator (i.e. Favilli et al., 2008). Assuming that both soils (alpine
456 permafrost and alpine non-permafrost) started to develop at the same time, the higher ages obtained in
457 the stable fraction of alpine permafrost soils are consequently reflecting a better stabilisation due to
458 permafrost conditions.
459 Contrary to the alpine sites, no difference in the vegetation type and cover was visible between
460 subalpine permafrost and subalpine non-permafrost soils (see chapter 2.1). In the Ah horizon of these
461 permafrost soils, higher carbon concentrations were found than in the respective horizon of the non-
462 permafrost soils. In addition, subalpine permafrost bulk soils showed a higher abundance of secondary
463 alcohols and polysaccharides and a decrease of the relative proportion in phenolic groups compared to
464 subalpine non-permafrost bulk soils (Fig. 4). Due to the higher amount of carbon recovered in the
465 lightest soil fractions (fPOM, oPOM) in subalpine permafrost soils, we suggest that subalpine
466 permafrost soils hold a higher amount of labile C while decomposition is more favoured in subalpine
467 non-permafrost soils. The assumption of a lower biodegradation in the subalpine permafrost bulk soil is
468 further supported by its high C/N ratio that indicates a higher amount of fresh or only weakly
469 decomposed organic matter (Sollins et al., 2006) compared to subalpine non-permafrost soils.
470 In the upper 15 cm of the subalpine soils, the ^{14}C-ages obtained in the subalpine bulk soils were similar
471 between permafrost and non-permafrost soils, whereas the stable C-fractions in permafrost soils were
472 younger compared to the non-permafrost site (Table 6). This was rather surprising and did not support

473 our hypothesis that older SOM should also be found in the permafrost soils at subalpine levels. The
474 reasons for this are still not clear. It appears that a better stabilisation of SOM with mineral surfaces or
475 inclusion into aggregates was achieved in the subalpine non-permafrost soils. These processes are
476 probably disrupted in the subalpine permafrost soils. From 15 to 30 cm soil depth, the subalpine
477 permafrost bulk SOM, however, showed higher ^{14}C-ages than the subalpine non-permafrost soil. This,
478 in turn, would support our hypothesis that decomposition processes are slower in permafrost soils.
479
480 *4.4. Climate warming and potential changes*
481 The response of decomposition and net-primary production (NPP) to climate warming is not balanced.
482 Decomposition rates are changing faster as a function of temperature than does net-primary production
483 (NPP) (Schimel et al., 1994). This imbalance in the temperature sensitivity between NPP and SOM
484 decomposition predicts that SOM storage should decrease in response to climatic warming (Lützow
485 and Kögel-Knabner, 2009), at least in the short term.
486 This is only in part supported by our results: we obtained higher ages for the bulk C and labile C-
487 fraction in permafrost soils (Tables 5 and 6; Fig. 6). Higher ages can be due to a lower input of C to the
488 soil and/or lower decomposition rates (that finally also depend on the communities of macro- and
489 micro-decomposer; Eskelinen et al., 2009; Seeber et al., 2009; Leifeld et al., 2012). In the stable C-
490 fraction, the ages follow, with some exceptions, better the 1:1 line (Fig. 6). The climate seems to be an
491 important driver for the bulk C and labile C-fraction (and less for the stable C-fraction).
492 In addition, differences in SOM quality (Fig. 4) and the higher contribution of potentially easily
493 degradable fPOM and oPOM fractions (lighter soil fractions, Fig. 5) to total SOM in permafrost show
494 that a climate warming could give rise to an increased microbial decomposition. However, the carbon
495 stocks in the bulk and labile fractions of the investigated permafrost soils were not significantly higher
496 than in non-permafrost soils. It is therefore questionable whether a thawing of permafrost really would
497 lead to an accelerated and increased carbon loss in these soils. For example, several scenarios of global

498 change are predicting ascending vegetation zones, with the subalpine coniferous forest and Ericaceous
499 shrubs becoming able to colonise meadows at higher altitudes (Ozenda and Borel, 1991). This shift in
500 vegetation is expected to increase primary production, but how such a shifting will influence the carbon
501 storage in the rooting zone is much more difficult to predict. Changes in plant species and a potential
502 increase in vegetation growth will enhance the aboveground carbon storage capacity that might offset
503 initial carbon losses.
504
505 **5. Conclusions**
506 We compared soils that have developed at sites either with or without permafrost conditions at a
507 subalpine and alpine level. We obtained the following main findings:
508 - Astonishingly, the organic carbon stocks did not significantly differ between permafrost and
509 non-permafrost sites. In addition the stocks were similar at the subalpine (forest) and alpine
510 (grassland) zone.
511 - Major differences between permafrost and non-permafrost influenced soils were however
512 detected in the chemical and physical characteristics of soil organic matter (SOM). The
513 chemical composition of SOM indicated differences in decomposition and stabilisation
514 mechanisms.
515 - The permafrost soils had a higher contribution of the free-particulate organic matter (fPOM)
516 and occluded organic matter (oPOM) to the total SOM. These fractions are potentially easily
517 degradable.
518 - In some permafrost soils a very high age of the stable SOM (resistant to H_2O_2 oxidation) could
519 be measured. At the alpine sites, higher ^{14}C ages of the bulk SOM and stable C-fraction reflect
520 lower SOM decomposition rates and a very long-term fixation of the stable C-fraction. At the
521 subalpine sites, rather opposite processes seem to occur that cannot as yet be explained.

522 The findings are not all equivocal and demonstrate that Alpine soils had a complex soil genesis that
523 now also affects their future behaviour (under the light of climate change). Carbon accumulation and C
524 fractionation occurred over a longer period (c. 5 – 8 ky BP) in a warmer climate that still imprints soil
525 characteristics to a certain degree. It is not really evident that the permafrost soils would release more
526 CO_2 with a warmer climate. Furthermore, a potential increased release might be offset due to the
527 shifting dominance from grassland to shrub/woodland.
528
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532
533 **References**
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Table 1
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Table 1. Chemical composition of the skeleton fraction of the parent material ('Julier Granite'; Böhlert et al., 2011a).

	SiO_2 g/kg	Al_2O_3 g/kg	Fe_2O_3 g/kg	CaO g/kg	MgO g/kg	K_2O g/kg	Na_2O g/kg	MnO g/kg	P_2O_5 g/kg	TiO_2 g/kg	ZrO_2 g/kg	LOI g/kg
Mean	686.3	148.8	30.9	22.9	9.6	28.8	29.8	0.60	0.98	4.09	0.20	18.3
SD	± 18.5	± 5.1	± 5.4	± 14.4	± 2.5	± 5.5	± 1.9	± 0.11	± 0.15	± 0.60	± 0.03	± 4.1

Table 2
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Table 2. Characteristics of the study sites.

Site	Coordinates[1] (N/E)	Elevation (m asl)	Aspect (°N)	Slope (%)	MAAT[2] (°C)	MAP[2] (mm/y)	Parent material	Vegetation	Soil orders (WRB)
Non-permafrost sites: Alpine									
Bever									
B1o	46°32'/9°48	2695	120	70	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Entic Podzol
B2o	46°32'/9°48	2680	120	78	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Haplic Regosol (humic)
B3o	46°32'/9°48	2677	120	36	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Entic Podzol
Albula									
A1o	46°35'/9°53	2615	150	70	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Entic Podzol
A2o	46°35'/9°53	2577	160	40	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Entic Podzol
A3o	46°35'/9°53	2569	160	42	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Entic Podzol
Non-permafrost sites: Subalpine									
Spinas									
S1o	46°22'/10°45'	1820	26	57	1.0	1050	Granite rich slope deposits	Natural forest (*Larici-Pinetum cembrae)*	Albic Podzol
S2o	46°22'/10°45'	1820	26	57	1.0	1050	Granite rich slope deposits	Natural forest (*Larici-Pinetum cembrae)*	Albic Podzol
S3o	46°33'/9°52	1796	40	46	1.0	1050	Granite rich slope deposits	Natural forest (*Larici-Pinetum cembrae)*	Haplic Podzol
Wo5, Wo6, Wo8, Wo9, Wo10	46°33'/9°15'	1733	40	46	1.0	1050	Granite rich slope deposits	Natural forest (*Larici-Pinetum cembrae)*	-
Permafrost sites: Alpine									
Bever									

B1m	46°32'/9°47’	2674	30	0	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Haplic Cambisol (Turbic)
B2m	46°32'/9°48’	2648	30	36	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Haplic Cambisol (Turbic)
B3m	46°32'/9°48’	2663	30	30	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Haplic Cambisol (Turbic)
Albula									
A1m	46°34'/9°53’	2620	70	62	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Entic Podzol
A2m	46°34'/9°53’	2620	55	62	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Entic Podzol
A3m	46°34'/9°53’	2616	50	70	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra *(Caricetum curvulae)*	Entic Podzol
Permafrost sites: Subalpine									
Spinas									
S1m	46°26'/9°57’	1820	20	57	1.0	1050	Granite rich slope deposits	Natural forest (*Larici-Pinetum cembrae)*	Haplic Cambisol (Dystric)
S2m	46°26'/9°57’	1820	20	57	1.0	1050	Granite rich slope deposits	Natural forest (*Larici-Pinetum cembrae)*	Haplic Cambisol (Dystric)
S3m	46°33'/9°51’	1808	40	46	1.0	1050	Granite rich slope deposits	Natural forest (*Larici-Pinetum cembrae)*	Haplic Cambisol (Dystric)
S4m	46°33'/9°51’	1809	40	53	1.0	1050	Granite rich slope deposits	Natural forest (*Larici-Pinetum cembrae)*	Haplic Podzol
Wm2, Wm3, Wm4, Wm7, Wm9	46°33'/9°51’	1719	40	46	1.0	1050	Granite rich slope deposits	Natural forest (*Larici-Pinetum cembrae)*	-

[1]World geodetic system (WGS84)
[2]MAAT = mean annual air temperature, MAP = mean annual precipitation (according to EDI, 1992; Leidlmair, 1996; Schwarb et al., 2000).
[3]WRB (IUSS Working Group 2006)

Table 3. Major IR absorption bands and assignments (Piccolo and Mirabella, 1985; Stevenson, 1994; Guo and Bustin, 1997; Senesi et al., 2003; Tan, 2003).

Band	Wave number cm^{-1}	Assignment
1	2980-2880	Aliphatic C-H stretching (aliphatic methyl and methylene groups)
2	1725-1710	C=O stretching of COOH, aldehydes and ketones
3	1660-1630	C=O stretching of amide groups, quinone C=O and/or C=O of H-bonded conjugated ketones
4	1620-1600	Aromatic C=C, strongly H-bonded C=O of conjugated ketones
5	1535-1500	Aromatic rings, amide II vibrations
6	1495-1470	N-H stretching of proteic amides
7	1470-1440	Aliphatic C-H stretching
8	1413-1333	OH deformation and C-O stretching of phenolic groups
9	1190-1127	C-OH stretching of aliphatic, alcoholic O-H
10	1116-1050	Secondary alcohols
11	1080-1030	C-O stretching of polysaccharide

Table 4
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Table 4. Physical and chemical characteristics including labile and stable organic carbon fractions (H_2O_2 treated samples) of the fine earth (<2mm).

Site	Horizon	Depth (cm)	Skeleton (wt.%)	Bulk density (g cm^{-3})	Sand (g kg^{-1})	Silt (g kg^{-1})	Clay (g kg^{-1})	pH ($CaCl_2$)	C_{tot} (g kg^{-1})	N_{tot} (g kg^{-1})	C_{tot}/ N_{tot}	C_{stable} (g kg^{-1})	N_{stable} (g kg^{-1})	C_{stable}/ N_{stable}	C_{labile} (g kg^{-1})	N_{labile} (g kg^{-1})	C_{labile} /N_{labile}	Crec[1)] %	Nrec[1)] %
Non-permafrost sites: Alpine																			
Bever																			
B1o	O	0-10	2	0.4	122	577	301	4.0	197	15	13	7.9	0.9	9	189.1	14.1	13	4.0	6.0
	OE	10-30	35	-	565	197	238	4.0	85.4	7.4	12	2.0	0.3	6	83.4	7.1	12	2.3	4.1
	Bs	30-50	54	1.25	760	214	26	4.3	25.9	3.4	8	2.7	2.7	1	23.2	0.6	38	10.4	79.4
	C	50-75	77	1.54	875	111	14	4.4	18.2	2.4	8	2.1	0	-	16.1	2.4	7	11.5	-
B2o	A	0-10	22	0.77	-	-	-	4.1	99.3	7.7	13	3.1	0.3	11	96.2	7.4	13	3.1	3.9
	AC	10-30	52	1.1	-	-	-	4.1	63.5	6.2	10	1.9	0.2	11	61.6	6.0	10	3.0	3.2
	bA	30-50	50	1.1	-	-	-	4.3	47.8	5.0	10	3.7	2.5	2	44.2	2.5	18	7.7	50.0
	R	>50	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
B3o	O	0-10	36	0.54	-	-	-	3.9	110.0	7.8	14	4.2	0.6	7	105.8	7.2	15	3.8	7.7
	AE	10-30	55	1.2	-	-	-	4.1	107.0	7.1	15	4.7	0.3	19	102.3	6.9	15	4.4	4.2
	Bhs	30-50	54	1.2	-	-	-	4.3	42.2	3.6	12	3.1	2.8	1	39.1	0.8	50	7.3	77.8
	Bs	50-60	49	1.37	-	-	-	4.5	13.7	2.3	6	2.2	0	-	11.5	2.3	5	16.1	-
	C	60-70	65	-	-	-	-	4.6	6.1	1.4	4	2.1	0	-	4	1.4	3	34.4	-
Albula																			
A1o	AE	0-10	19	0.75	483	341	176	3.8	33.0	3.0	11	2.5	1.9	1	30.5	1.1	28	7.6	63.3
	BA	10-20	32	1.1	448	371	180	4.1	15.9	2.3	7	2.6	2.1	1	13.3	0.3	53	16.4	91.3
	Bhs1	20-40	49	1.1	485	345	170	4.1	15.0	2.1	7	2.5	1.5	2	12.5	0.5	23	16.7	71.4
	Bhs2	40-60	60	1.1	723	224	52	4.4	27.9	1.9	15	3.4	0	-	24.5	1.9	13	12.2	-
	C	60-70	49	-	653	305	42	4.5	10.9	0.2	50	2.5	0	-	8.4	0.2	38	22.9	-
A2o	AE	0-10	50	0.74	-	-	-	3.8	73.9	5.5	13	3.6	2.6	1	70.3	2.9	24	4.9	47.3
	Bhs1	10-20	52	1	-	-	-	4.4	39.8	4.1	10	2.6	2.1	1	37.2	2.0	19	6.5	51.2
	Bhs2	20-35	56	1	-	-	-	4.2	14.8	2.8	5	2.3	1.8	1	12.5	1.0	12	15.5	64.3
	C	35-45	68	-	-	-	-	4.5	8.7	1.6	5	2.9	0	-	5.8	1.6	4	33.3	-
A3o	A	0-10	17	0.85	-	-	-	3.5	79.6	7.4	11	3.5	2.6	1	76.1	4.8	16	4.4	35.1
	AE	10-20	46	0.85	-	-	-	3.8	43.6	5.1	9	3.9	2.1	2	39.7	3.0	13	8.9	41.2

	Bh1	20-40	65	1.07	-	-	-	4.2	32.3	3.6	9	2.7	2.1	1	29.6	1.6	19	8.4	58.3
	Bh2	40-60	62	1.07	-	-	-	4.3	29.2	3.1	9	2.8	3	1	26.4	0.2	168	9.6	96.8
	C	60-75	60	-	-	-	-	4.5	13.5	2.5	5	2.4	0	-	11	2.5	5	17.8	-
Non-permafrost sites: Subalpine																			
Spinas																			
S1o	E	0-10	69	1.19	728	205	67	4.0	58.4	3.2	18	5.6	1.0	6	52.7	2.3	23	9.6	31.3
	Bs1	10-20	57	-	750	214	36	3.4	49.9	3.8	13	5.2	0.7	7	44.7	3.1	15	10.4	18.4
	Bs2	20-30	68	1.18	749	220	31	4.3	37.7	2.2	18	5.0	1.3	4	32.6	0.8	39	13.3	59.1
	Bs3	30-40	60	-	778	184	38	4.4	35.1	2.7	13	3.5	0	-	31.6	2.7	12	10.0	-
	B	50-60	58	1.13	649	233	118	4.4	31.9	2.7	12	3.8	2.6	2	28.1	0.2	166	11.9	96.3
	R	>60	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
S2o	E	0-10	36	-	-	-	-	3.8	63.5	5.8	11	3.7	1.0	4	59.8	4.8	12	5.8	17.2
	Bs1	10-20	40	-	-	-	-	4.2	31.0	3.4	9	3.1	0.8	4	27.9	2.7	11	10.0	23.5
	Bs2	40-50	54	-	-	-	-	4.5	19.8	2.3	9	2.9	2.3	1	16.9	0	0	14.6	100.0
	B	70-80	82	-	-	-	-	4.4	15.8	2.4	7	2.5	0	-	13.3	2.4	6	15.8	-
	R	>80	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
S3o	A	0-10	63	0.51	-	-	-	3.8	65.7	3.2	21	7.4	1.0	8	58.3	2.2	26	11.3	31.3
	E	10-20	54	0.51	-	-	-	3.8	91.5	3.7	25	9.4	0.7	13	82.1	3.0	28	10.3	18.9
	Bh1	20-30	83	0.93	-	-	-	3.7	151.5	4.5	34	11.0	1.9	6	140.5	2.6	55	7.3	42.2
	Bh2	30-40	59	0.93	-	-	-	3.7	130.0	4.1	32	8.2	2.0	4	121.8	2.1	58	6.3	48.8
	IIE	40-50	41	0.79	-	-	-	3.7	25.3	1.2	21	3.0	0	-	22.3	1.2	19	11.9	-
	Bs	60-100	67	-	-	-	-	4.6	29.2	0.7	42	3.8	0	-	25.4	0.7	36	13.0	-
	R	>100	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Wo5	A	0-15	28	0.55	-	-	-		71.7	6.6	11	5.6	2.1	3	66	4.5	15	7.8	31.8
	B	15-30	50	1.22	-	-	-		19.3	2.6	7	2.5	0	-	16.8	2.6	6	13.0	-
Wo6	A	0-15	38	0.55	-	-	-		81	5.9	14	5.2	1.9	3	75.8	4.0	19	6.4	32.2
	B	15-30	27	1.22	-	-	-		47.1	3.6	13	3.6	1.9	2	43.4	1.7	26	7.6	52.8
Wo8	A	0-15	48	0.55	-	-	-		31.6	4.1	8	3.3	2.0	2	28.3	2.1	13	10.4	48.8
	B	15-30	48	1.22	-	-	-		22.8	3.2	7	3.2	1.9	2	19.6	1.3	15	14.0	59.4

Wo9	A	0-15	18	0.55	-	-	-		146.5	9.0	16	9.0	1.4	7	137.5	7.6	18	6.1	15.6
	B	15-30	21	1.22	-	-	-		52.1	4.5	12	2.4	2.5	1	49.7	2.0	25	4.6	55.6
Wo10	A	0-10	40	0.55	-	-	-		128.0	8.5	15	11.2	1.8	6	116.8	6.7	18	8.8	21.2
	B	15-20	33	1.22	-	-	-		37.8	3.9	10	3.1	1.0	6	34.8	3.4	10	8.2	25.6
Permafrost sites: Alpine																			
Bever																			
B1m	A	0-10	38	1.56	793	130	77	4.5	15.8	3.1	5	1.2	0.1	17	14.6	3.0	5	7.6	3.2
	Bw1	10-30	39	1.56	848	89	63	4.5	12.8	2.2	6	0.9	0.1	13	11.9	2.2	6	7.0	4.5
	Bw2	30-50	40	1.57	786	133	81	4.5	20.1	2.6	8	2.4	0	-	17.7	2.6	7	11.9	-
	BC	50-60	52	1.67	814	103	83	4.6	11.6	2.1	6	2.3	0	-	9.3	2.1	4	19.8	-
	R	>60	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
B2m	Ah	0-10	31	1.34	-	-	-	4.6	19.6	3.1	6	1.2	0.1	13	18.4	3.0	6	6.1	3.2
	Bw1	10-30	26	1.34	-	-	-	4.6	12.1	2.6	5	2.0	0.1	35	10.0	2.5	4	16.5	3.8
	Bw2	30-35	18	1.36	-	-	-	4.6	12.6	1.9	7	2.6	0	-	10.0	1.9	5	20.6	-
	Bw3	35-50	26	1.33	-	-	-	4.6	14.9	2.5	6	2.4	0	-	12.5	2.5	5	16.1	-
	C	50-80	32	1.74	-	-	-	4.6	12.7	2.3	5	2.2	0	-	10.5	2.3	4	17.3	-
B3m	Ah	0-10	29	1.11	-	-	-	4.4	29.6	3.8	8	2.5	1.4	2	27	2.4	11	8.4	36.8
	Bw	10-30	35	1.27	-	-	-	4.6	11.5	2.7	4	2.4	1.9	1	9	0.8	11	20.9	70.4
	BC1	30-50	42	1.61	-	-	-	4.7	7.2	1.7	4	2.0	0	-	5.2	1.7	3	27.8	-
	BC2	50-60	43	1.61	-	-	-	4.7	6.1	2.0	3	1.6	0	-	4.5	2.0	2	26.2	-
	R	>60	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Albula																			
A1m	A1	0-10	31	1.15	543	293	164	3.9	27.4	2.3	12	2.5	1.7	2	24.9	0.6	45	9.1	73.9
	A2	10-20	34	1.29	512	332	156	4.0	20.8	1.8	12	2.9	1.7	2	17.9	0.2	118	13.9	94.4
	AE	20-40	32	1.18	524	313	163	4.1	21.4	1.3	16	2.7	0	-	18.7	1.3	14	12.6	-
	Bs	40-80	54	1.42	619	272	109	4.6	22.1	1.2	18	3.3	0	-	18.8	1.2	16	14.9	-
	BC	80-100	67	1.62	674	277	49	4.3	6.6	0.3	26	2.4	0	-	4.1	0.3	17	36.4	-
	R	>100	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
A2m	AE1	0-10	48	0.78	-	-	-	3.6	71.2	4.1	18	2.8	2.4	1	68.4	1.7	41	3.9	58.5
	AE2	10-20	68	0.93	-	-	-	3.9	42.2	3.8	11	2.8	1.4	2	39.3	2.5	16	6.6	36.8
	Bhs	20-45	55	1.33	-	-	-	4.3	20.4	1.7	12	2.3	1.4	2	18.1	0.4	49	11.3	82.4

	B	45-80	62	1.57	-	-	-	4.5	10.3	2.4	4	2.4	0	-	7.9	2.4	3	23.3	-
	C	80-100	44	1.42	-	-	-	4.6	3.6	0.2	17	2.3	0	-	1.3	0.2	6	63.9	-
A3m	A	0-10	62	0.66	-	-	-	3.8	96.4	5.6	17	4.6	2.2	2	91.8	3.5	27	4.8	39.3
	AE1	10-20	63	0.66	-	-	-	3.9	63.6	3.8	17	3.1	2.2	1	60.5	1.6	39	4.9	57.9
	AE2	20-35	67	0.66	-	-	-	4.1	55.6	2.7	21	3.3	2.1	2	52.2	0.6	90	5.9	77.8
	Bhs	35-60	59	1.20	-	-	-	4.5	32.3	1.4	23	2.6	0	-	29.7	1.4	21	8.0	-
	C	60-70	65	1.66	-	-	-	4.6	9.8	0.2	42	2.2	0	-	7.6	0.2	33	22.4	-
Permafrost sites: Subalpine																			
Spinas																			
S1m	A	0-10	55	1.15	633	218	149	3.8	40.0	4.3	9	3.8	1.1	4	36.2	3.3	11	9.5	25.6
	Bw	25-35	73	1.2	611	239	150	4.4	21.3	2.5	9	2.6	1.7	2	18.7	0.8	22	12.2	68.0
	BC	50-60	51	1.39	622	234	144	4.5	21.3	2.4	9	2.6	0	-	18.7	2.4	8	12.2	-
	R	>60	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
S2m	A	0-10	50	-	-	-	-	3.9	64.4	3.5	18	3.3	1.2	3	61.1	2.4	26	5.1	34.3
	Bw	25-35	61	-	-	-	-	4.1	45.2	2.2	21	3.8	1.2	3	41.4	1.0	43	8.4	54.5
	BC	50-60	51	-	-	-	-	4.1	49.0	2.4	20	3.8	0	-	45.2	2.4	19	7.8	-
	R	>60	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
S3m	Ah	0-10	52	0.65	-	-	-	3.5	93.2	5.5	17	5.5	1.2	5	87.7	4.3	20	5.9	21.8
	Bw1	10-20	53	0.84	-	-	-	4.0	36.6	3.0	12	2.8	0.8	4	33.8	2.2	15	7.7	26.7
	Bw2	20-30	73	0.84	-	-	-	4.1	32.6	3.0	11	3.1	1.3	2	29.5	1.7	17	9.5	43.3
	BC1	30-40	75	0.84	-	-	-	4.2	33.2	2.8	12	2.9	2.8	1	30.3	0	0	8.7	100.0
	BC2	40-50	68	0.84	-	-	-	4.3	29.3	2.1	14	3.7	0	-	25.6	2.1	12	12.6	
	C	50-60	60	-	-	-	-	4.3	33.1	3	11	3.6	2.3	2	29.5	0.7	46	10.9	76.7
S4m	AE	0-10	53	0.48	-	-	-	3.9	57.4	2.1	27	5.2	1.1	5	52.2	1.0	53	9.1	52.4
	Bh	10-20	59	0.48	-	-	-	3.9	35.8	0.8	45	4.9	0	-	30.9	0.8	39	13.7	-
	Bhs1	20-30	54	0.82	-	-	-	4.2	34.3	1.1	31	4.0	0	-	30.3	1.1	28	11.7	-
	Bhs2	30-40	62	0.73	-	-	-	4.3	39.9	0.7	57	3.8	0	-	36.1	0.7	52	9.5	-
	Bhs3	40-50	54	0.73	-	-	-	4.3	33.0	0.2	165	4.1	0	-	28.9	0.2	145	12.4	-
	Bhs4	50-60	64	0.73	-	-	-	4.3	40.5	0.8	51	5.2	0.2	32	35.3	0.6	55	12.8	25.0
	Bs1	60-70	62	-	-	-	-	4.3	36.8	0.8	46	3.7	0.2	23	33.1	0.6	52	10.1	25.0
	Bs2	70-90	73	-	-	-	-	4.3	20.9	0.6	35	2.6	0.2	14	18.3	0.4	44	12.4	33.3

	R	>90	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Wm2	A	0-10	16	0.16	-	-	-		200.0	11.2	18	18.2	1.3	14	181.9	10.0	18	9.1	11.6
	B	10-20	35	0.54	-	-	-		78.0	4.8	16	7.6	1.8	4	70.4	3.0	24	9.7	37.5
Wm3	A	0-15	34	0.16	-	-	-		280.5	14.0	20	19.2	1.3	15	261.4	12.7	21	6.8	9.3
	B	15-30	19	0.54	-	-	-		117.0	7.3	16	9.7	2.0	5	107.4	5.4	20	8.3	27.4
Wm4	A	0-15	14	0.16	-	-	-		354.5	13.8	26	25.3	0.7	34	329.2	13.0	25	7.1	5.1
	B	15-25	30	0.54	-	-	-		246.0	11.0	23	15.6	1.3	12	230.4	9.6	24	6.3	11.8
Wm7	A	0-15	25	0.16	-	-	-		146.0	7.8	19	6.7	1.9	4	139.3	5.9	24	4.6	24.4
	B	15-30	27	0.54	-	-	-		123.0	7.5	17	8.9	1.7	5	114.1	5.7	20	7.2	22.7
Wm9	A	0-15	38	0.16	-	-	-		195.5	11.2	17	12.6	1.4	9	182.9	9.8	19	6.4	12.5
	B	15-25	63	0.54	-	-	-		34.7	3.6	10	2.6	1.8	1	32.0	1.7	18	7.5	50.0

[1)] Recovery values

Table 5
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Table 5. Radiocarbon dating (obtained using ^{14}C) of the bulk soil, the labile and stable organic carbon fractions (H_2O_2 treated samples) of the alpine sites (permafrost and non-permafrost).

Site	Depth (cm)	bulk C				labile C-fraction			stable C-fraction			
		pMC	Uncalibrated ages (y uncal. BP)	$\partial^{13}C$ (‰)	Calibrated ages (y cal. BP) *2 sigma range*	pMC	Uncalibrated ages (y uncal. BP)	Calibrated ages (y cal. BP) *2 sigma range*	pMC	Uncalibrated ages (y uncal. BP)	$\partial^{13}C$ (‰)	Calibrated ages (y cal. BP) *2 sigma range*
non-permafrost												
B1o	0-10	91	675 ± 30	-25.3 ± 1.1	631 - 679	92	640	562 - 591	82	1570 ± 30	-26.4 ± 1.1	1394 – 1530
	10-30	89	920 ± 25	-21.6 ± 1.1	777 - 920	89	890	760 - 799	75	2305 ± 30	-30.2 ± 1.1	2305 – 2357
B2o	0-10		modern	-	modern	-	-	modern		875 ± 35	-28.5 ± 1.1	725 – 835
	10-30	92	595 ± 35	-26.0 ± 1.1	538 - 654	93	551	532 - 555	76	2135 ± 35	-25.8 ± 1.1	2000 – 2162
B3o	0-10	86	1170 ± 30	-26.8 ± 1.1	1049 - 1175	86	1117	979 - 1057	71	2645 ± 30	-23.7 ± 1.1	2733 – 2796
	10-30	79	1815 ± 30	-26.0 ± 1.1	1691 - 1825	80	1743	1613 - 1700	64	3555 ± 30	-29.4 ± 1.1	3817 – 3927
permafrost												
B1m	0-10	76	2170 ± 30	-26.2 ± 1.1	2106 - 2311	76	2109	2041 - 2130	69	2980 ± 35	-24.7 ± 1.1	3061 – 3268
	10-30	68	3015 ± 40	-19.8 ± 1.1	3078 - 3341	68	3003	3161 - 3256	67	3180 ± 40	-25.7 ± 1.1	3335 – 3479
B2m	0-10	93	485 ± 30	-30.3 ± 1.1	499 - 545	95	394	469 - 501	77	2025 ± 35	-33.1 ± 1.1	1892 – 2064
	10-30	73	2495 ± 35	-22.1 ± 1.1	2456 - 2736	81	1613	1509 - 1543	31	9465 ± 35	-33.6 ± 1.1	10584 - 10788
B3m	0-10	83	1450 ± 30	-26.1 ± 1.1	1299 - 1389	84	1335	1271 - 1294	70	2785 ± 30	-29.7 ± 1.1	2838 – 2957
	10-30	68	3000 ± 30	-22.9 ± 1.1	3078 - 3269	71	2704	2766 - 2845	59	4205 ± 30	-30.8 ± 1.1	4627 - 4763

Table 6

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Table 6. Radiocarbon dating (obtained using ^{14}C) of the bulk soil, the labile and stable organic carbon fractions (H_2O_2 treated samples) of the subalpine sites (permafrost and non-permafrost).

Site	Depth (cm)	bulk C				labile C-fraction			stable C-fraction			
		pMC	Uncalibrated ages (y uncal. BP)	$\partial^{13}C$ (‰)	Calibrated ages (y cal. BP) *2 sigma range*	pMC	Uncalibrated ages (y uncal. BP)	Calibrated ages (y cal. BP) *2 sigma range*	pMC	Uncalibrated ages (y uncal. BP)	$\partial^{13}C$ (‰)	Calibrated ages (y cal. BP) *2 sigma range*
non-permafrost												
Wo5	0-15	94	430 ± 25	-28.1 ± 1.1	461 - 525	95	359	434 - 480	84	1315 ± 25	-26.8 ± 1.1	1228 – 1295
						-	-	-		-	-	-
Wo6	0-15	100	-35 ± 25	-27.9 ± 1.1	41 - 60	102	modern	modern	72	2590 ± 25	-29.7 ± 1.1	2711 – 2760
	15-30		55 ± 25	-27.8 ± 1.1	31 - 84	-	-	-		-	-	-
Wo8	0-15	106	-500 ± 25	-28.2 ± 1.1	46 - 55	110	modern	modern	66	3295 ± 25	-28.5 ± 1.1	3453 – 3579
	15-30		modern	-27.8 ± 1.1	modern	-	-	-		-	-	-
Wo9	0-15	106	-560 ± 25	-27.9 ± 1.1	46 - 55	107	modern	modern	90	775 ± 25	-30.2 ± 1.1	673 – 731
	15-30		modern	-30.4 ± 1.1	modern	-	-	-		-	-	-
permafrost												
Wm2	0-15	102	-235 ± 25	-27.2 ± 1.1	46 - 55	104	modern	modern	89	855 ± 25	-29.6 ± 1.1	695 – 796
	15-30		400 ± 30	-30.6 ± 1.1	428 - 514	-	-	-		-	-	-
Wm3	0-15	108	-655 ± 25	-27.2 ± 1.1	46 - 54	109	modern	modern	96	310 ± 25	-29.6 ± 1.1	348 – 459
	15-30		modern	-24.2 ± 1.1	modern	-	-	-		-	-	-
Wm4	0-15	104	-400 ± 25	-28.2 ± 1.1	46 - 54	106	modern	modern	84	1295 ± 25	-30.6 ± 1.1	1178 – 1287
	15-30		145 ± 25	-23.5 ± 1.1	185 - 224	-	-	-		-	-	-

Figure captions

Fig. 1. Above: Study sites at the Val Bever and the Albula pass with the approximate locations of the sampling sites: (1) Bever, alpine; (2) Albula, alpine: (3) Spinas, subalpine. Below: Photographs of the sampling sites and their characteristic soil profiles. (1) alpine site: 'non-permafrost' (left) and 'permafrost' (right). (3) subalpine site: no difference in the terrain between both sites and typical soil profiles; 'non-permafrost' (left) and 'permafrost' (right).

Fig. 2. A) Carbon recovery after the H_2O_2 treatment for permafrost and non-permafrost soils. Alpine sites (Bever, Albula): left half of the plot; Subalpine site (Spinas): right half of the plot. B) C/N ratios of the bulk soil (untreated) and stable (H_2O_2-treated) C-fraction of permafrost and non-permafrost soils. Alpine sites (Bever, Albula): left half of the plot; subalpine site (Spinas): right half of the plot.

Fig. 3. Stocks of total, labile and stable (recovery after the H_2O_2-treatment) organic carbon in permafrost and non-permafrost soils. Alpine sites (Bever, Albula): left; subalpine site (Spinas): right. Carbon stocks were calculated down to the C-horizon or rock surface.

Fig. 4. Box plots (median, 25th- and 75th-percentile values, min. and max. values) of the relative proportion of A) secondary alcohols, B) C-O stretching of polysaccharide, C) OH deformation and C–O stretching of phenolic groups and D) N-H stretching of proteic amides of the bulk soil (untreated) and stable (H_2O_2-treated) C-fractions of permafrost and non-permafrost soils. Alpine sites (Bever, Albula): left; subalpine site (Spinas): right.

Fig. 5. Carbon concentration in the different density fractions in the topsoil (0-15cm) of the subalpine site (N=3). Fractions of oPOM and mOM where additionally treated with an ultrasonic energy of 50 J ml^{-1}.

Fig. 6. Comparison of the SOC age range of paired samples (permafrost vs. non-permafrost sites and corresponding soil horizons). A) bulk C, B) labile C, C) stable C, D) stable C including additional subalpine sites (in red: Egli et al., 2009). Age cold soils = age range of the permafrost soils (this investigation) and additional subalpine (north-facing) sites and corresponding soil horizons (Egli et al., 2009). Age less cold soils = age range of the non-permafrost soils (this investigation) and additional subalpine (south-facing) sites and corresponding soil horizons (Egli et al., 2009).

Figure 1

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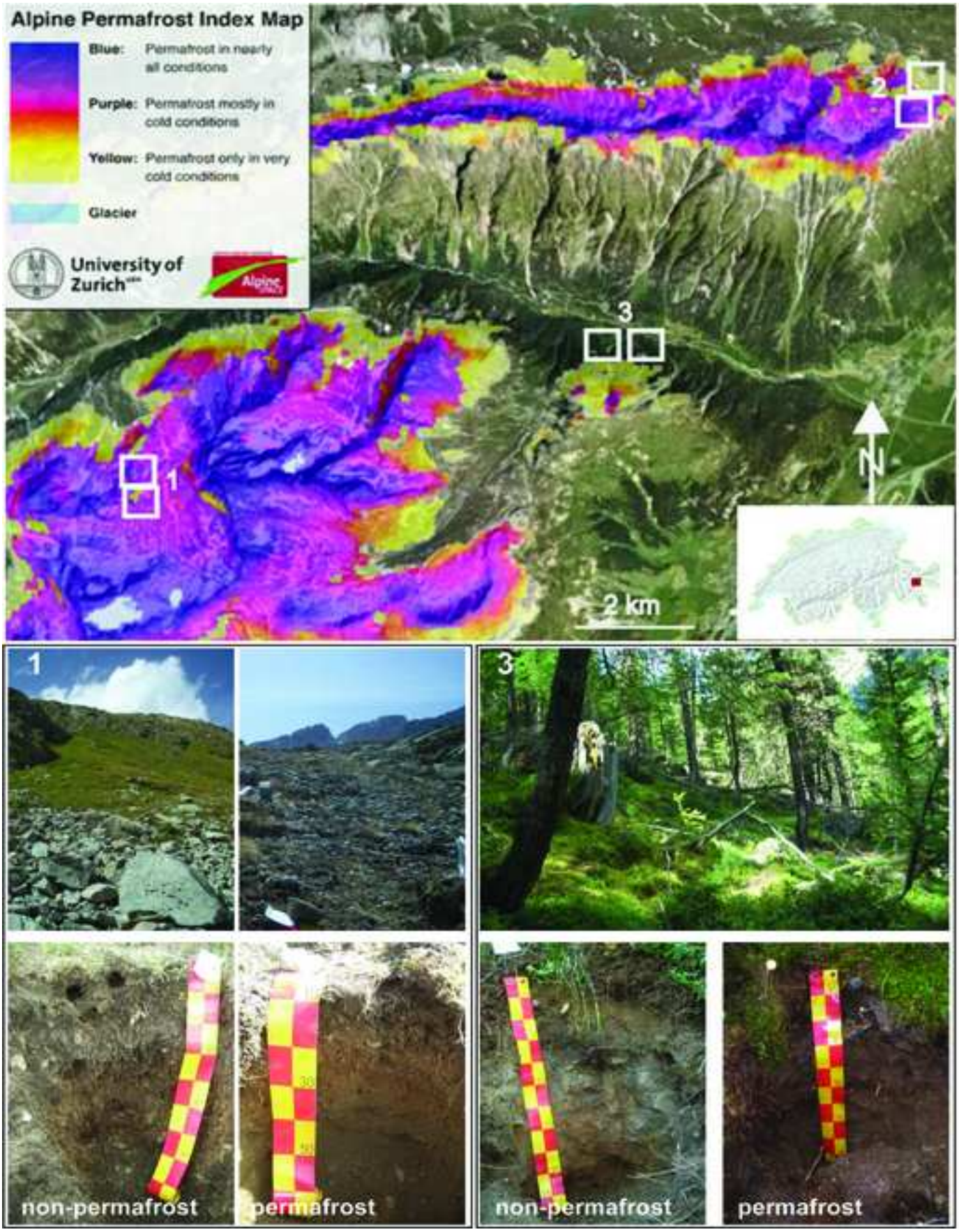

Figure 2

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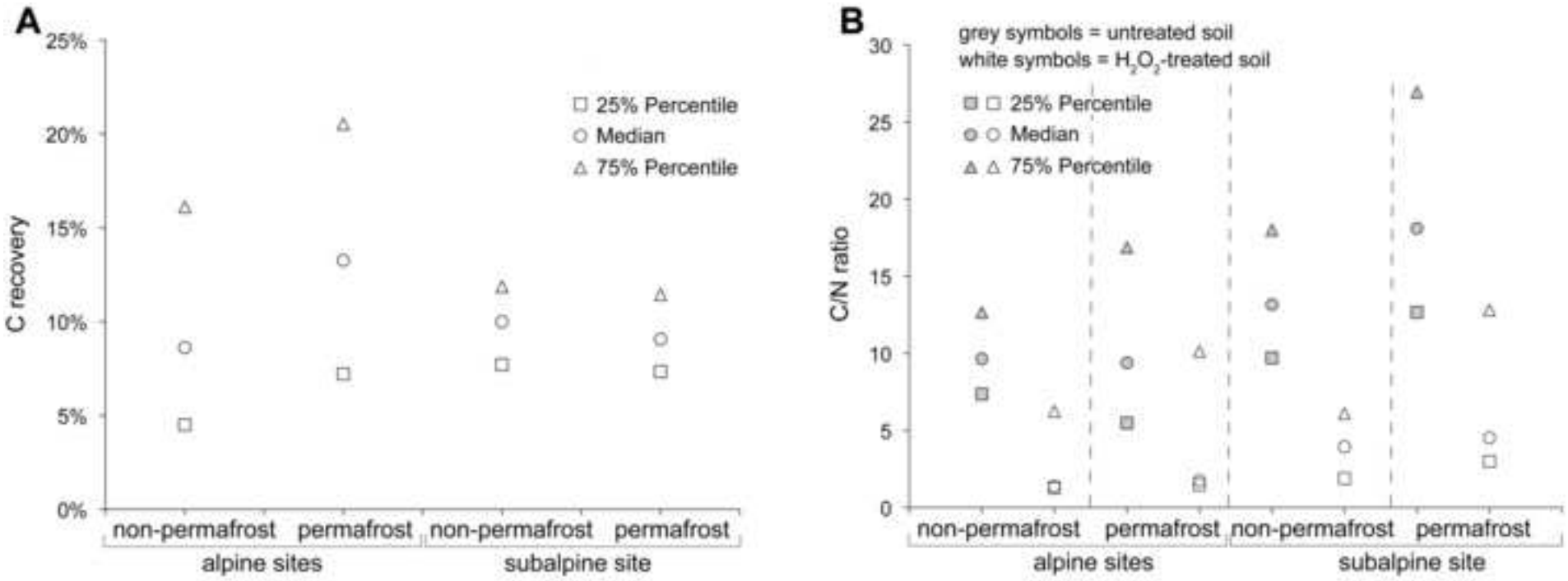

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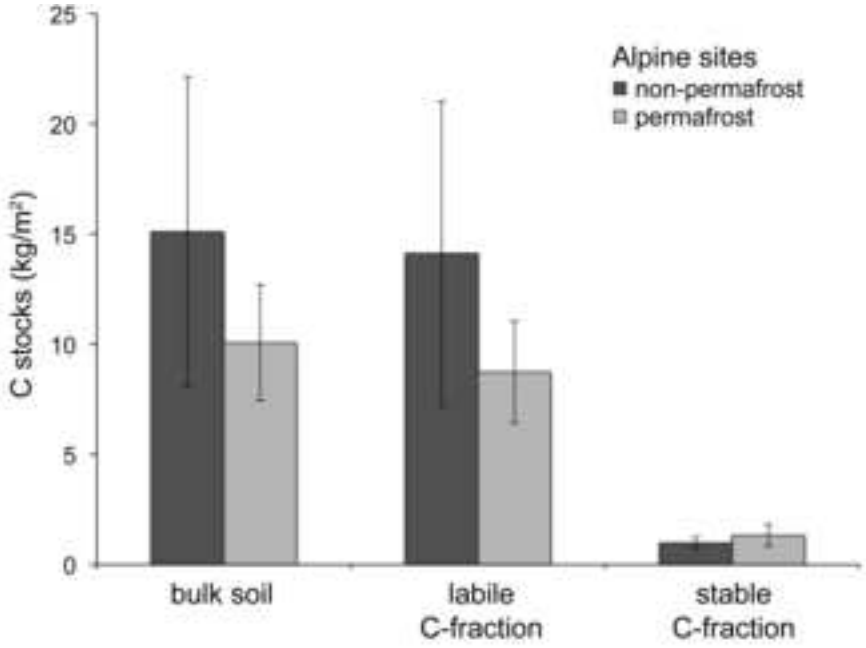

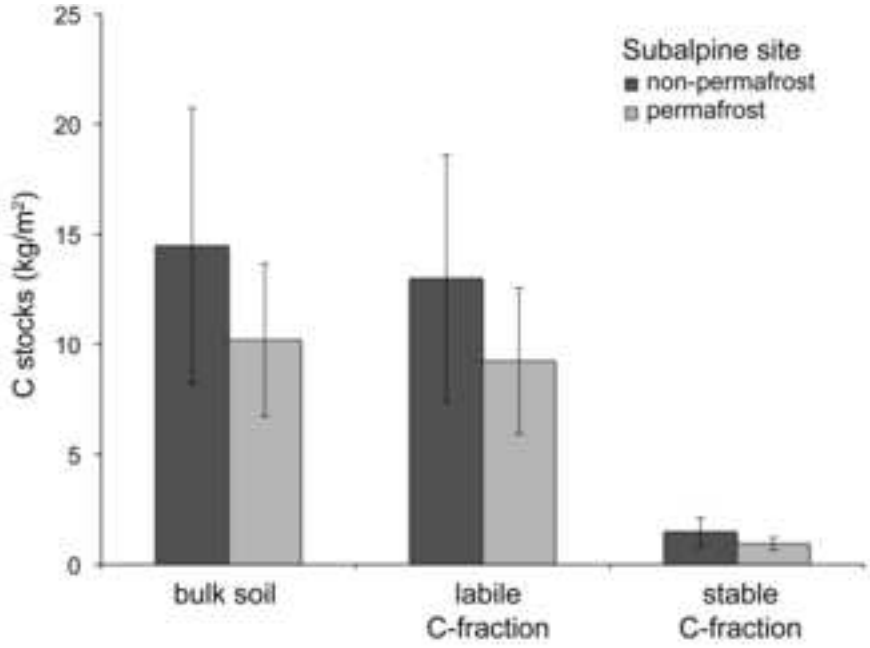

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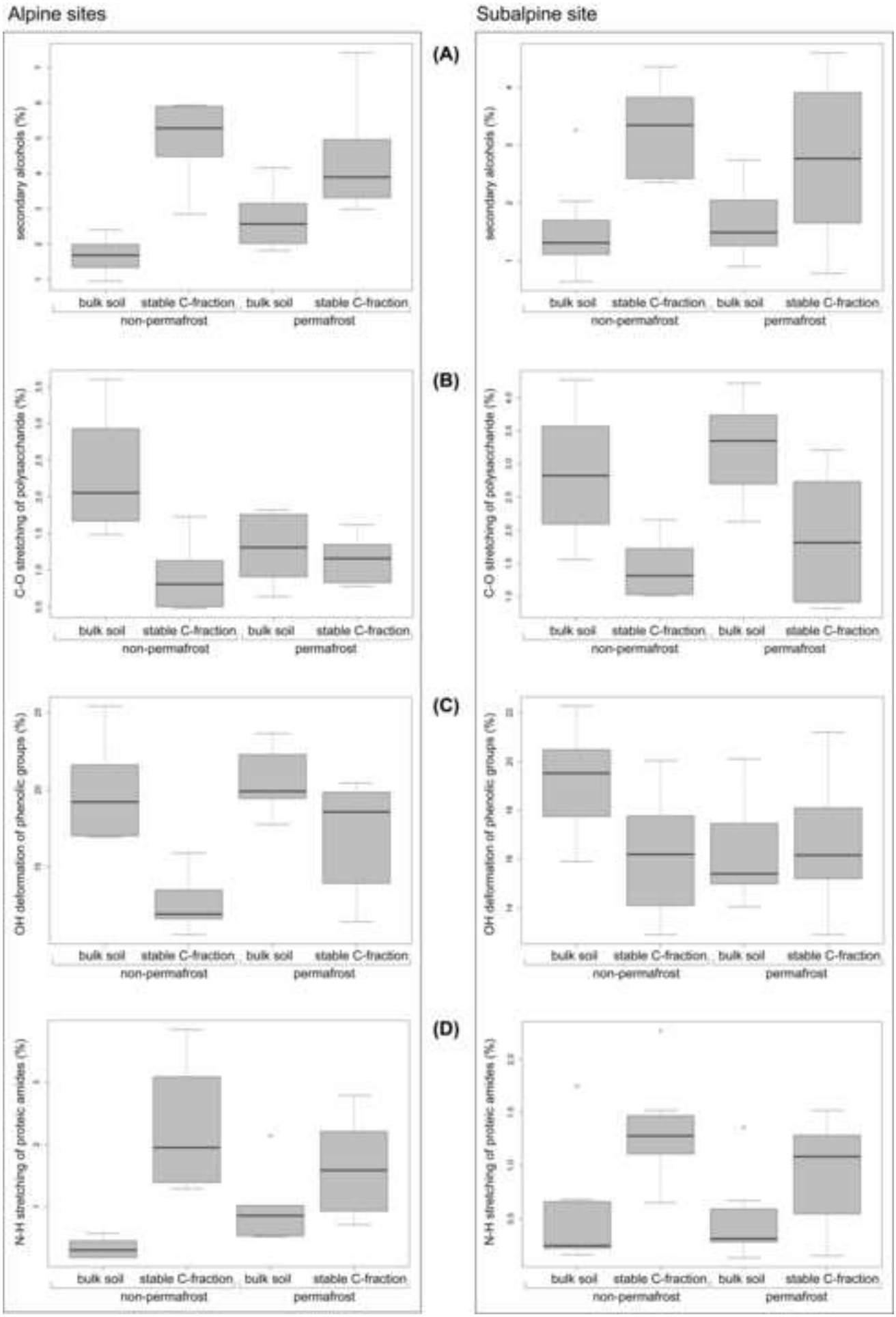

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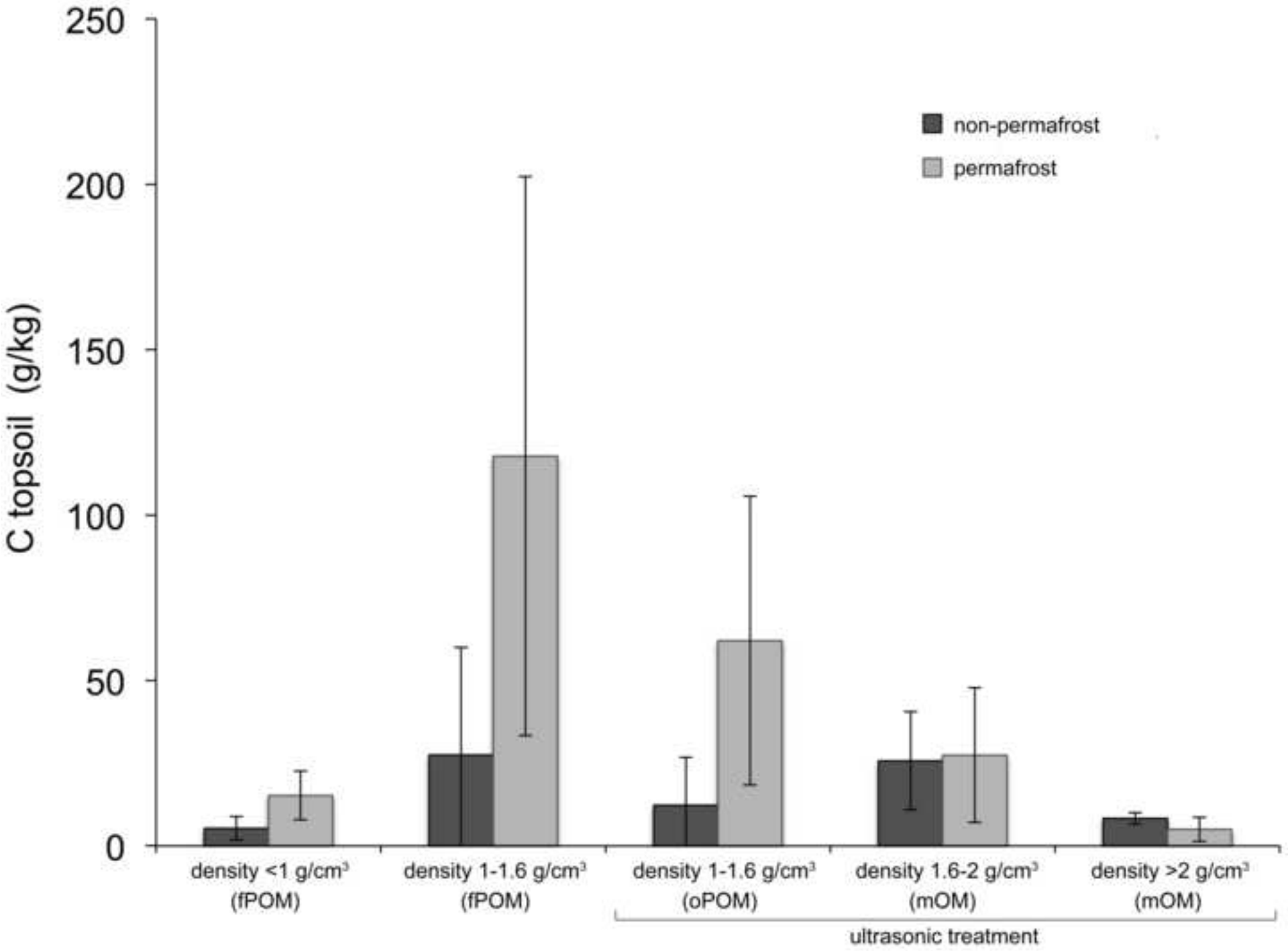

Figure 6

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